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Data synthesis of carbon distribution in particle size fractions of tropical soils: Implications for soil carbon storage potential in croplands

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ABSTRACT

Organic carbon saturation in soils refers to the theoretical maximum soil organic carbon (SOC) that can be associated with and stabilized on fine silt plus clay particles ($F < 20\ \mu m$). We reviewed the literature dealing with SOC distribution between soil fractions to evaluate carbon saturation for tropical soils and estimate the C storage potential of cropland.

We collected 258 data points on SOC distribution between soil fractions in tropical soils from 84 sites in 27 countries. We used boundary line analysis to estimate the maximum stabilized SOC depending on soil group, clay type and land use. SOC storage potential was calculated as the SOC saturation deficit, the difference between the actual SOC content and the maximum stabilized SOC content.

We found that the maximum SOC in the fine fraction of tropical soils ($53\ g\ C\ kg^{-1}$ fine fraction) was lower than previous assessments of global SOC storage scale based mainly on temperate soils. The $F < 20\ \mu m$ fractions were closer to SOC saturation in forest soils than in croplands. The cropland had a higher soil C storage potential, but changing agricultural management practices did not fill the deficit that is calculated using the whole dataset. The deficit was much lower when it was estimated with grassland or cropland data only: this provides a more realistic estimate for SOC storage potential for croplands.

The SOC content in the coarser fraction ($F > 50\ \mu m$) did not depend on soil texture and significantly contributed to the total SOC, especially in sandy soils (41.3%). This is affected by changes in agricultural management practices. We concluded that, although the aim of increasing SOC stabilization originally arose from climate change mitigation strategies, it must now be more viewed as being more relevant to food security and local adaptation to climate change.

1. Introduction

Soil organic carbon (SOC) is the largest carbon sink in the terrestrial biosphere, amounting to about 1500 Pg for the top meter of soil (excluding Cryosols) (Jobbágy and Jackson, 2000). SOC plays a fundamental role in the fertility and productivity of terrestrial ecosystems, supporting important soil-derived ecosystem services such as soil quality, water filtration, erosion control, nutrient cycling, habitat and energy for soil organisms (Lal, 2016; Milne et al., 2015). Soil carbon storage has also been recognized as an efficient option to offset the rise in global atmospheric CO_2 concentration (Smith, 2016). Thus, storing carbon in soil is important for both food security and climate mitigation and adaptation.

Aboveground and belowground plant litter is the primary source of SOC. SOC accumulates in soils through the decomposition and transformation of litter by soil organisms. The formation of organo-mineral complexes in the fine silt- and clay-size fractions ($F < 20\ \mu m$) has been recognized as the most important process for SOC stabilization (Feller and Beare, 1997; Hassink, 1997; Six et al., 2002a), far more important than chemical recalcitrance of the organic matter (Dungait et al., 2012; Marschner et al., 2008). Hassink (1997) used least-squares linear regression of the relationship between SOC associated with the fine particle fraction, expressed in $g\ C$ in $F < 20\ \mu m\ kg^{-1}$ soil, and the relative mass of this fraction, expressed in $g\ kg^{-1}$ soil, to estimate the capacity of the fine particles to stabilize carbon in uncultivated and grassland soils. The slope of this linear regression represented the theoretical capacity of the fine fraction of soil to accumulate and stabilize SOC. Further reviews studying the relationship between these variables highlighted the influence of the clay type (2:1 versus 1:1) and land use (i.e. forests, grasslands, croplands) on the slope of the

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regression (Feng et al., 2013; Six et al., 2002a). Feng et al. (2013) showed that the use of least-squares regression underestimated the maximum SOC that could be stabilized in the fine particle fraction and suggested the use of boundary line analysis to estimate the maximum stabilized SOC.

SOC saturation based on the maximum amount of SOC that can be stabilized on fine particles has been used in the literature to estimate the SOC storage potential, as the deficit between theoretical maximum stabilized SOC content for a given soil sample and the measured SOC content. Some studies (Angers et al., 2011; Wiesmeier et al., 2014) have estimated the SOC storage potential at national scale using the saturation limit as defined by Hassink (1997) or using fine particle content and other soil predictors like mineral surface area, exchangeable Al, Fe, and pH (Beare et al., 2014; McNally et al., 2017). These approaches address long-term, i.e. decadal, storage potential but do not take into account the short-term, i.e. annual, benefits of particulate organic matter. The size of this pool is highly variable and depends on land use and climate (Gregorich et al., 2006; Wang et al., 2016; Wiesmeier et al., 2014). Another way to estimate the SOC storage potential is to calculate the difference between measured SOC stocks or content in pristine conditions and measured SOC stocks in cultivated or human-impacted conditions in areas with the same soil and climate conditions (Akpa et al., 2016). This method also depends on the concept of SOC saturation as it assumes that soils in pristine conditions are higher in SOC than in anthropized areas due to higher C inputs, which is challenged by some studies dealing with pastures (e.g. Fujisaki et al., 2015; Stahl et al., 2016). A second assumption is the saturated state of soils in SOC under pristine vegetation. However, this latter assumption might overestimate the SOC storage potential, e.g. in the case of cultivated lands that remain cultivated. Stewart et al. (2007) introduced the concept of "effective stabilization capacity", where the upper limit of SOC storage depends not only on physical and chemical properties of the soil but also on the levels of soil disturbance that vary depending on land use and within agroecosystems (e.g. tillage). Soils under native vegetation would, therefore, have a greater SOC stabilization capacity than cultivated soils.

SOC saturation and storage potential has not been investigated recently in tropical soils, despite its importance. The most recent meta-analysis of SOC distribution in the fine particle fraction was at global scale (Feng et al., 2013). However, tropical soils have specific chemical and mineralogical properties that could have a different effect on SOC saturation levels than those of temperate soils (Barthès et al., 2008; Feller and Beare, 1997). In recent years tropical soils have been subjected to land-use changes to increase production of food, timber, and fibers (Grace et al., 2014; Hansen et al., 2013). Large areas in the tropics have also been degraded and eroded (Gibbs and Salmon, 2015; Kiage, 2013), so we could expect low SOC stocks in these areas and thus a large potential for SOC storage.

To our knowledge the capacity of tropical cultivated soils to increase SOC content in the fine particle fraction to the SOC saturation level has not yet been evaluated. Is the SOC storage potential calculated using SOC saturation an achievable goal for increasing SOC to contribute to climate change mitigation and food security?

The general objective of this study was to review the literature to evaluate SOC storage potential in the tropics using the distribution of SOC in soil fractions. We aimed to i) define the level of SOC saturation in fine particle fractions depending on the soil type, clay type, and land use; ii) evaluate the potential for reducing the SOC saturation deficit by improving practices for annual crops; iii) assess the contribution of particulate organic matter to SOC content.

2. Materials and methods

2.1. Data collection

Data were collected by searching existing peer-reviewed literature supplemented by searches for relevant grey literature. We searched literature published up to 2016 dealing with SOC distribution in particle-size fractions. The literature search was restricted to studies covering any area between the tropics or having a tropical climate according to the IPCC climate classification based on elevation, mean annual temperature and precipitation (IPCC, 2010, 2006). Google Scholar query «soil carbon "particle-size fractions" "20 µm" tropic*» produced 705 results. Only English language search terms were used but a few articles and PhD dissertations in French or Portuguese were also considered. We also checked papers that cited methodological papers dealing with this topic (Christensen, 1992; Gavinelli et al., 1995) and the papers cited in Feng et al. (2013) which is the most recent review dealing with this topic globally.

We selected studies where SOC content was measured in the fine ($F < 20$ µm) and/or coarse-size ($F > 50$ µm) fractions of the soil. SOC content associated to one of these fractions and expressed in g C in fraction kg^{-1} soil is calculated through Eq. (1).

$$SOC_F = F \times C_F \tag{1}$$

where SOC_F is the SOC associated to the considered fraction (g C in fraction kg^{-1} soil), F is the relative mass of the fraction (g fraction kg^{-1} soil), and C_F is the SOC content of the fraction (g C kg^{-1} fraction). Depending on the results presented in the collected studies, we could directly extract SOC_F values, or calculated them with Eq. (1). In addition some studies used separate analyses of the fractions < 2 µm and 2–20 µm. In this case, SOC_F of these fractions were summed to obtain the SOC associated to the $F < 20$ µm.

We included all studies using particle-size fractionation to separate organic matter depending on soil particle-size classes whatever the methods applied to disperse soil aggregates: these could use glass beads, ultrasonication, hexametaphosphate pre-treatment, Na resin, or combination of these techniques as described in Christensen (1992) and Gavinelli et al. (1995). In order to limit the dispersion in the dataset, we excluded studies using low energy dispersion fractionation methods aiming to generate aggregates. The recovery rate of soil fractionation was not always provided so we did not use a filter for this criterion. However, observations with a SOC recovery rate (sum of SOC across the fractions/bulk SOC content) higher than 1.2 were discarded. To produce the quantitative overview of the SOC content in the particle-size fractions, results from topsoil layers, i.e. up to 10 cm depth, were used. Some studies analyzed 0–5 and 5–10 cm layers separately, in this case we averaged the two values to obtain SOC content and particle-size mass for the 0–10 cm layer. The selection process gave 45 papers, 8 of which were included in the review by Feng et al. (2013), giving 258 observations in 84 sites and 27 countries in tropical areas. Most of the observations were located in South America and Sub-Saharan Africa. We found a few studies in Asia and Oceania (8 data points, see the map in Supplementary Fig. 1).

Soil type, clay type, and land use were collected from each paper (Supplementary Table 1). When not directly provided in the paper, we assigned soil to a WRB reference soil group (IUSS Working Group WRB, 2015) based on the soil classes in other classification systems, soil properties and/or any relevant information reported in each paper. Then, we further assembled these reference groups into four major groups according to the group sequences described in the Soil Atlas of Africa (Jones et al., 2013). These groups are based on the dominant factors or processes that most clearly control the formation of the soil. Group I included relatively homogeneous sandy and young soils with limited or poor profile development, such as Arenosol (24 data points in the collected dataset) and Cambisol (13 data points). Group II included soils with a clay-rich or argic subsoil horizon with a low base saturation, low activity clay (Acrisol, 47 data points), high base saturation, high activity clay (Luvisol, 8 data points) or high base saturation, low activity clay (Lixisol, 43 data points). Group III included soils where iron and/or aluminum chemistry plays a major role in their formation, mainly Ferralsols (98 data points) and few Nitisols (4 data points).

Andosols derived from volcanic ash and dominated by short-range-order mineral or Al-humus complexes belong in this group. However, stabilization of SOC in Andosols not only results from SOC adsorption on amorphous or poorly crystalline mineral surfaces (Beare et al., 2014) but also from the poor accessibility of SOC trapped in the mesopore structure that may create technical problems for particle-size fractionation (Chevallier et al., 2010). For these reasons, and as we found only two studies dealing with Andosols, we did not include them in this study. Group IV included soils that are or have been strongly influenced by water, Vertisols (19 data points) and Planosols (2 data points).

Additionally, as different clays have different capacities for binding and stabilizing SOC (Greenland, 1965; Six et al., 2002a), the clay type for each observation was classified as 1:1 (218 data points) or 2:1 (40 data points), based on the identification of the clay type in the original paper or by inference from the soil reference groups. Finally, the dataset was classified by land use: forest ($n = 49$), grassland ($n = 92$), and cropland ($n = 117$). Forest included natural forests and tree plantations, with a relatively high tree density, maintained for timber, non-timber cash or food production including fruit- or nut-producing trees in orchards. Grassland included natural grasslands and savannas, grazed and mowed pastures, and herbaceous fallows. Cropland included annual crops, perennial crops (mainly sugarcane), agroforestry systems with an annual cropping and low-tree density, abandoned fields after cropping, and slash-and-burn systems.

2.2. Data analysis

We analyzed the relationship between the SOC associated with the fine-size fraction (g C in F $<$ 20 μm kg^{-1} soil) as the dependent variable and the relative mass of this fraction in the soil (g F $<$ 20 μm kg^{-1} soil) as the independent variable. We evaluated this relationship with i) least-squares linear regression (LR) (Hassink, 1997) and ii) boundary line analysis ($BL_{0.9}$), with quantile regression (R quantreg package, Koenker, 2016) applied to the upper decile of the data ($\tau = 0.9$). As many soils in the dataset are supposed to be far from saturation (Feng et al., 2013), the boundary line analysis should allow the evaluation of the maximum SOC stabilization in the fine particle fraction, i.e. the SOC saturation limit (C_{sat}). Linear and quantile regressions were all forced through the origin because any positive intercept would represent SOC that is not stabilized in fine soil particles (Feng et al., 2013). We checked that the intercepts were not significantly different from 0 at $P < 0.05$ when the regressions were performed without forcing the intercept. We performed LR and $BL_{0.9}$ on the whole dataset and for each land use. We calculated R^1 to evaluate the local goodness of fit of quantile regression curves (Koenker and Machado, 1999). The slopes of the least-squares regressions were compared using the post-hoc Tukey test at $P < 0.05$.

As measurements of the SOC in different fractions are still rare, we also tested if there was a relationship between the total SOC content and the relative mass of F $<$ 20 μm in our dataset, this relationship being called "textural control of SOC retention" by Zinn et al. (2007). This was assessed with least-squares linear regression and quantile regression on the upper decile, but without using forced zero intercepts as the coarse soil fractions may contribute to the SOC. Additionally, we used quantile regression on the first decile ($\tau = 0.1$) in order to assess the lower limit of SOC stabilization ($BL_{0.1}$).

We also evaluated the relationship between the SOC associated with the $>$ 50 μm fraction (F $>$ 50 μm), corresponding to the particulate organic matter (Christensen, 2001), and the relative mass of the F $>$ 50 μm. F $>$ 50 μm was preferred to F $>$ 20 μm because we focused on particulate organic matter fraction. This fraction is generally defined as the particulate carbon found in the fraction 50–2000 μm (e.g. Gregorich et al., 2006). In addition, a large number of studies that presented results for SOC associated with F $>$ 50 μm did not analyze the fraction 20–50 μm of $>$ 20 μm (Supplementary Table 1). Taking into account only the F $>$ 20 μm for the particulate carbon would reduce significantly the number of observations of the synthesis. The SOC associated with the fraction $>$ 50 μm and the contribution of the fraction $>$ 50 μm to the total SOC content, i.e. the ratio between SOC in the fraction $>$ 50 μm and the total SOC, were compared between soil groups and land uses with Kruskal-Wallis tests ($P < 0.05$) as ANOVA assumptions were not met.

In order to assess the SOC saturation deficit and changes of SOC content in soil fractions for cultivated soils, we selected data from 9 papers in our dataset (35 data points in 13 sites) that compared a control cropland with agricultural systems implementing practices that are expected to improve SOC content such as manuring, mineral amendment, crop diversification, tillage reduction, and residue retention. We also compared cropland with grassland at the same sites.

The maximum SOC amount stabilized in fine particles, C_{sat} (g C kg^{-1} soil), was calculated using Eq. (2).

$$C_{sat} = BL_{0.9} \times F < 20\ \mu m \quad (2)$$

where $BL_{0.9}$ is the slope of the upper boundary line of the relationship between SOC associated with F $<$ 20 μm and the relative mass of F $<$ 20 μm, and F $<$ 20 μm the relative mass of the fraction $<$ 20 μm (g fraction kg^{-1} soil). We calculated the $BL_{0.9}$ of the whole dataset and for forest, cropland or grassland only. From this, we calculated SOC saturation levels (or effective SOC stabilization capacity) depending on the land use.

The SOC saturation deficit, $C_{sat\text{-}def}$, was calculated using Eq. (3).

$$C_{sat-def} = C_{sat} - SOC_{F<20\ \mu m} \quad (3)$$

where C_{sat} is the SOC saturation level (g C kg^{-1} soil) and $SOC_{F<20\ \mu m}$ the measured SOC content associated with F $<$ 20 μm (g C kg^{-1} soil). $C_{sat\text{-}def}$ was calculated for the whole dataset as well as for sub datasets to estimate an SOC saturation level (effective C stabilization capacity) for particular land uses ($C_{sat\text{-}defLU}$).

We also calculated the differences in SOC in the F $<$ 20 μm and in the coarse fraction (F $>$ 20 μm) between control cropland and the modified agroecosystems. We chose to report SOC changes in F $>$ 20 μm rather than F $>$ 50 μm in order to take account the whole SOC variations after cropland management changes.

Statistical analyses were carried out using R software (R Core Team, 2016).

3. Results

3.1. SOC associated with the < 20 μm fraction

We found a positive relationship between the relative mass of F $<$ 20 μm and SOC content associated with this fraction ($y = 0.033 \pm 0.001\,x$ for the linear regression) (Fig. 1): on average the F $<$ 20 μm held 33 g C kg^{-1} F $<$ 20 μm.

SOC content associated with F $<$ 20 μm (g C kg^{-1} soil) was lowest for Arenosols and Cambisols in soil group I. There was no clear distinction between soil groups II and III, i.e. soils with clayey subsoils (mainly Acrisols) and soils distinguished by Fe–Al chemistry (mainly Ferralsols). Except for few cases, the SOC content associated with F $<$ 20 μm for group IV soils, mostly Vertisols with 2:1 clays, were below the linear regression line of the whole dataset (Fig. 1b).

The results showed that the slope of $BL_{0.9}$ (53 ± 7 g C kg^{-1} F $<$ 20 μm) was higher than LR (33 ± 1 g C kg^{-1} F $<$ 20 μm) for the whole dataset (Fig. 2). Very few cropland soil samples seemed to reach the maximum stabilized SOC content calculated as $BL_{0.9}$ for the whole dataset. The slope of the linear regression in croplands was about half the slope of the $BL_{0.9}$ for the whole dataset (25 ± 1 and 53 ± 7 g C kg^{-1} F $<$ 20 μm, respectively). Some forest and grassland soil samples however seemed to reach the calculated maximum stabilized SOC content (Fig. 2). The linear regression calculated only for forest soils was similar to the $BL_{0.9}$ calculated for the whole dataset. However, there was a large variability of SOC content associated with

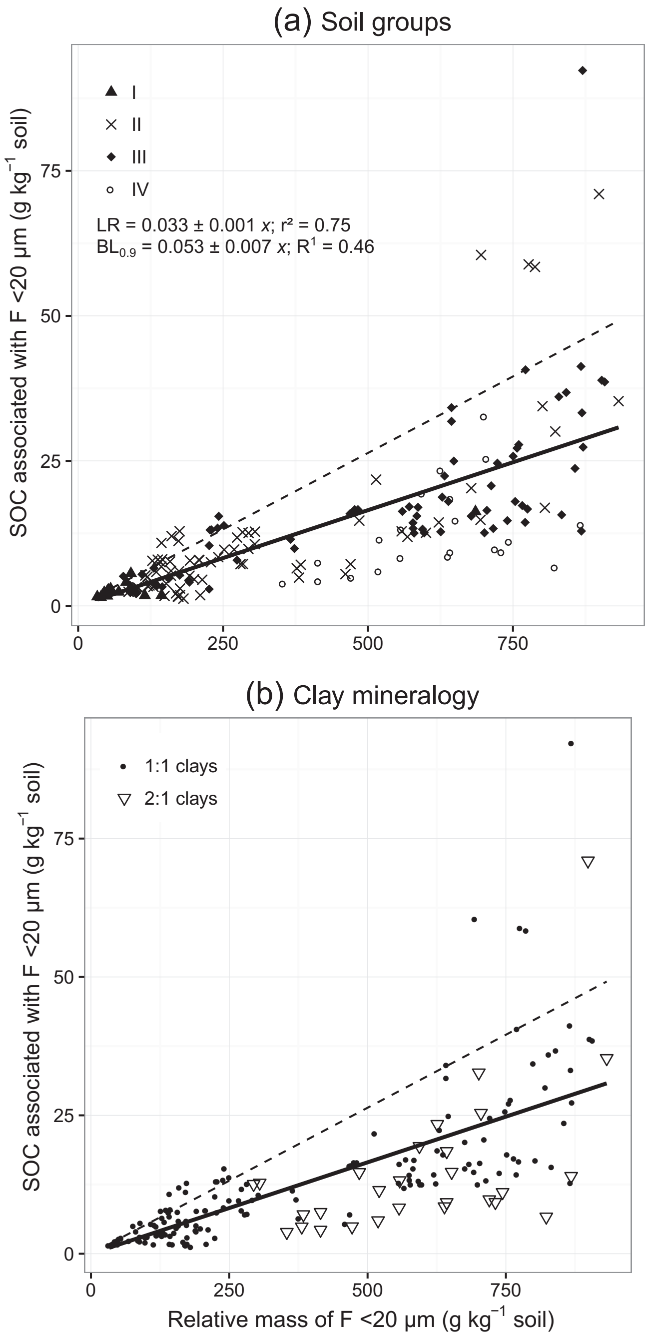

Fig. 1. Relationship between SOC associated with F < 20 μm (g C in F < 20 μm kg^{-1} soil) and relative mass of F < 20 μm (g F < 20 μm kg^{-1} soil) depending on (a) soil groups and (b) clay type. Soil group I includes Arenosols and Cambisols, group II includes Luvisols, Acrisols and Lixisols, group III includes Ferralsols and Nitisols and group IV includes Vertisols and Planosols. The solid lines are the linear regression lines (LR) through the origin (n = 178). The dashed lines are the upper boundary line ($BL_{0.9}$) through the origin for the whole dataset.

F < 20 μm in forest soils, especially in soils with a high fine particle content.

Land use affected the slopes of the linear regressions and the $BL_{0.9}$ (upper boundary line). The calculation of $BL_{0.9}$ for each land use showed that the maximum stabilized SOC associated with F < 20 μm was 87 ± 12 g C kg^{-1} F < 20 μm under forest, 47 ± 6 g C kg^{-1} soil under grassland, and 38 ± 4 g C kg^{-1} F < 20 μm under cropland. The slope of the linear regressions was significantly higher under forest (57 ± 5 g C kg^{-1}) than under grassland and cropland (34 ± 1 and 25 ± 1 g C kg^{-1} respectively). Even for soils with low fine particle content, the SOC associated with F < 20 μm was lower in grassland and cropland than in forest soils.

3.2. SOC associated with the F > 50 μm fraction

There was no significant relationship between the SOC content associated with F > 50 μm and the relative mass of F > 50 μm in the soil (Fig. 3).

The average contribution of F > 50 μm to the total SOC content was 25.9 ± 15.1% (n = 195). The contribution of F > 50 μm was the highest for Arenosols and Cambisols, from soil group I (Table 1). Within the other soil groups, SOC contents in F > 50 μm were higher in grasslands than in croplands.

3.3. Total SOC content

We found a pattern of textural control of SOC retention, as there was a positive relationship between total SOC content and the mass of the F < 20 μm (Fig. 4). The slope of the linear regression was significantly higher for forest (68 ± 1 g C kg^{-1} F < 20 μm) than grassland and cropland soils (43 ± 3 and 30 ± 3 g C kg^{-1} F < 20 μm respectively) at P < 0.05. Overall, the slopes of the linear regressions were higher than those found for the relationship between SOC content in F < 20 μm and the mass of F < 20 μm (Figs. 2 and 4). The intercepts were positive but not significantly different from zero at P < 0.05.

The data dispersion was higher for forest soils than grassland and cropland soils, as shown by area between the upper and lower boundary lines (Fig. 4).

3.4. SOC saturation and changes in SOC content with changes in cropland management

In the collected dataset, a limited number of papers (9 out of 45) compared control croplands and agroecosystems implementing practices that are expected to increase SOC content. It encompassed a wide range of F < 20 μm mass and SOC saturation values (Table 2). The relative mass of F < 20 μm varied from 123 to 823 g kg^{-1} soil whereas the SOC saturation ranged from 6.5 g C in F < 20 μm kg^{-1} soil to 43.6 g C in F < 20 μm kg^{-1} soil for soils with the largest clay content (Table 2). The lowest SOC saturation deficit values were found for sandy soils: around 3–4 g C kg^{-1} in Senegal and Togo (Feller et al., 1991). The SOC saturation deficit generally increased with the proportion of fine particles and reached a maximum of 29.4 g C kg^{-1} in a bare Vertisol in Martinique (French West Indies).

Overall, the conversion of cropland into grasslands caused larger increases in SOC than changes in management practices for annual crops. The largest increase for annual crops was obtained for improved farming practices on a chromic Luvisol in Tanzania (Solomon et al., 2000). In this case, the SOC saturation deficit was 3.4 g C kg^{-1} in soils of smallholding fields where animal manure was regularly applied for about 10 years followed by 5 years of bare fallow against 15.3 g C kg^{-1} for fields cultivated without any inputs (Table 2). In some studies, with medium- or fine-textured soils, changes in management practices for cropland increased the SOC associated with F < 20 μm by up to 7.9 g C in F < 20 μm kg^{-1} soil, bringing the soils closer to SOC saturation (Table 2 see Grandière et al., 2007; Razafimbelo et al., 2006; Solomon et al., 2000).On the contrary, other studies did not show an increase in SOC content in F < 20 μm (de Freitas et al., 2000; Metay et al., 2007). For sandy soils, the changes in SOC content in F < 20 μm with changes in cropland management practices were moderate, from − 0.3 to + 1.7 g C in F < 20 μm kg^{-1} soil, and without a clear trend while the increases in SOC were the highest in the > 20 μm fraction, especially after manure has been added (Table 2 see Feller et al., 1991; Hien, 2004). For Vertisols, the case studies showed that the SOC saturation deficit decreased from 21.2 down to 9.8 g C kg^{-1} soil several years after replacing market gardening by grassland (Larré-Larrouy

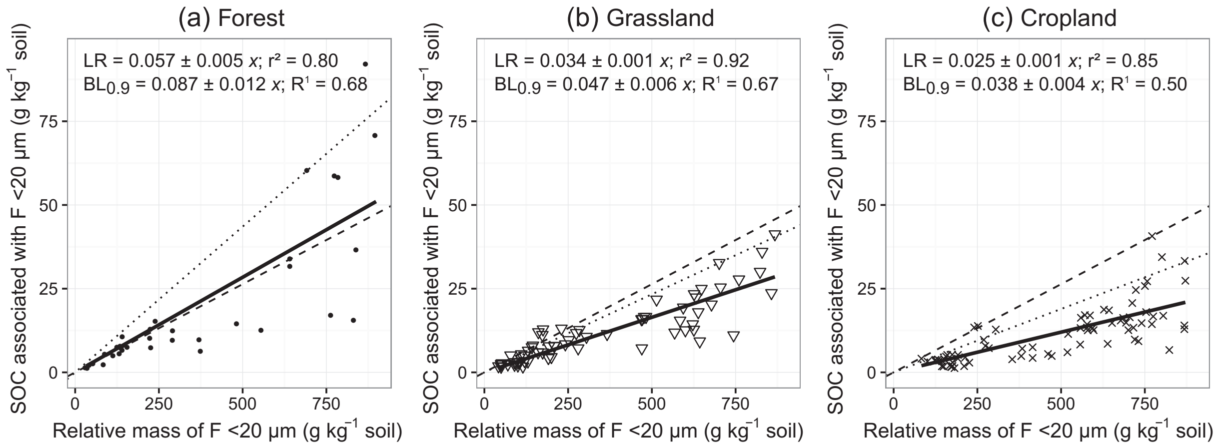

Fig. 2. Relationship between SOC content associated with F $<$ 20 µm and relative mass of F $<$ 20 µm depending on land use (n = 34 for forest, n = 66 for grassland, n = 78 for cropland). The solid lines are the linear regression lines (LR) through the origin for each land use. The dotted lines are the upper boundary lines ($BL_{0.9}$) through the origin for each land use (significantly higher for forest than grassland and cropland at $P < 0.05$). The dashed lines are the upper boundary lines through the origin for the whole dataset, also shown in Fig. 1 (slope = 0.053).

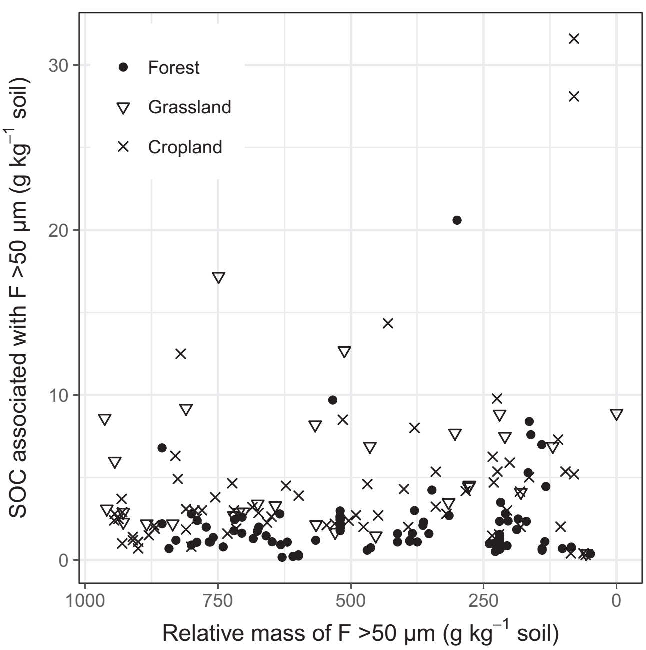

Fig. 3. Relationship between SOC content associated with F $>$ 50 µm and relative mass of F $>$ 50 µm depending on land use.

et al., 2003) and the increase in SOC was mostly in F $<$ 20 µm, being up to 15 g C kg^{-1} soil (Table 2; Chevallier, 1999; Feller et al., 2001; Larré-Larrouy et al., 2003).

The SOC saturation deficits were lower when using $BL_{0.9}$ calculated for cropland or grassland only ($C_{sat\text{-}defLU}$). For example, in Burkina Faso, at Saria, sorghum fields with mineral and organic amendment had a SOC saturation deficit $C_{sat\text{-}defLU}$ of 3.7–3.9 g C in F $<$ 20 µm kg^{-1} soil against a global SOC saturation deficit $C_{sat\text{-}def}$ of 6.2–6.3 g C in F $<$ 20 µm kg^{-1} soil (Table 2 see Hien, 2004). For medium-textured manured smallholding fields studied by Solomon et al. (2000), the SOC saturation deficit $C_{sat\text{-}defLU}$ had negative value of −1.2 g C in F $<$ 20 µm kg^{-1} soil (Table 2). Soil under annual crops converted to no tillage with cover crops in Madagascar almost reached the SOC saturation as calculated for croplands only (Grandière et al., 2007). The SOC saturation deficits for Vertisols, 10 years after reversion to grassland were calculated using the $C_{satGRASSLAND}$ boundary line and, therefore only moderately decreased (Table 2 see Feller et al., 2001; Larré-Larrouy et al., 2003; Chevallier, 1999).

Table 1
SOC associated with F $>$ 50 µm and its relative contribution to total SOC content depending on soil group and land use. Means are followed by standard errors. Means with the same upper case letter do not significantly differ between groups (Kruskal-Wallis test, $P < 0.05$). Means with the same lower case letters do not significantly differ between land uses for a given soil group (Kruskal-Wallis test, $P < 0.05$). Soil group I includes to Arenosols and Cambisols, soil group II includes Luvisols, Acrisols and Lixisols, soil group III includes Ferralsols and Nitisols, and soil group IV includes Vertisols and Planosols.

Soil group	Land use	SOC in F > 50 µm (g kg^{-1} soil)	SOC in F > 50 µm/total SOC (%)	n
All	All	3.4 ± 4.0	25.9 ± 15.1	195
All	Forest	4.76 ± 3.7 (a)	25.7 ± 16.2	35
	Grassland	4.3 ± 5.1 (a)	27.6 ± 13.6	70
	Cropland	2.3 ± 2.6 (b)	24.6 ± 15.6	90
I	All	4.3 ± 3.5 (A)	41.3 ± 15.4 (A)	34
	Forest	5.4 ± 3.8	45.6 ± 19.7	9
	Grassland	4.2 ± 3.4	40.6 ± 12.8	17
	Cropland	3.3 ± 3.3	38.6 ± 16.3	8
II	All	2.4 ± 1.8 (B)	28.1 ± 13.0 (B)	73
	Forest	3.6 ± 3.0 (ab)	23.9 ± 11.7	10
	Grassland	3.1 ± 2.0 (a)	26.0 ± 7.5	23
	Cropland	1.8 ± 0.8 (b)	31.2 ± 15.6	40
III	All	3.8 ± 5.3 (AB)	18.4 ± 12.5 (C)	74
	Forest	5.1 ± 4.1 (a)	18.6 ± 11.0 (ab)	16
	Grassland	5.3 ± 7.9 (a)	22.9 ± 14.3 (a)	25
	Cropland	2.1 ± 2.1 (b)	14.4 ± 10.4 (b)	33
IV	All	4.5 ± 5.0 (AB)	25.1 ± 11.4 (BC)	14
	Grassland	5.1 ± 1.0 (a)	26.5 ± 15.3	5
	Cropland	4.1 ± 6.3 (b)	24.5 ± 10.1	9

4. Discussion

4.1. SOC associated with the < 20 µm fraction

The maximum stabilized SOC of 53 g C kg^{-1} fine fraction for the whole dataset (n = 178) was lower than the value of 78 g C kg^{-1} fine fraction found by Feng et al. (2013) for all climate zones (n = 342). In both studies, the maximum stabilized SOC content was calculated using a boundary line analysis on datasets with a variety of soil types and land uses. In Feng et al. (2013), most of the observations that determined the SOC saturation line were from 2:1 clay temperate soils under grassland or forest. Differences in climate could explain the lower SOC saturation line in our case. Higher SOC turnover rates in tropical climates could result in lower SOC stabilization in the fine particle fraction of tropical soils (Six et al., 2002b). However, the effect of climate on the maximum stabilized SOC content in fine fractions remains

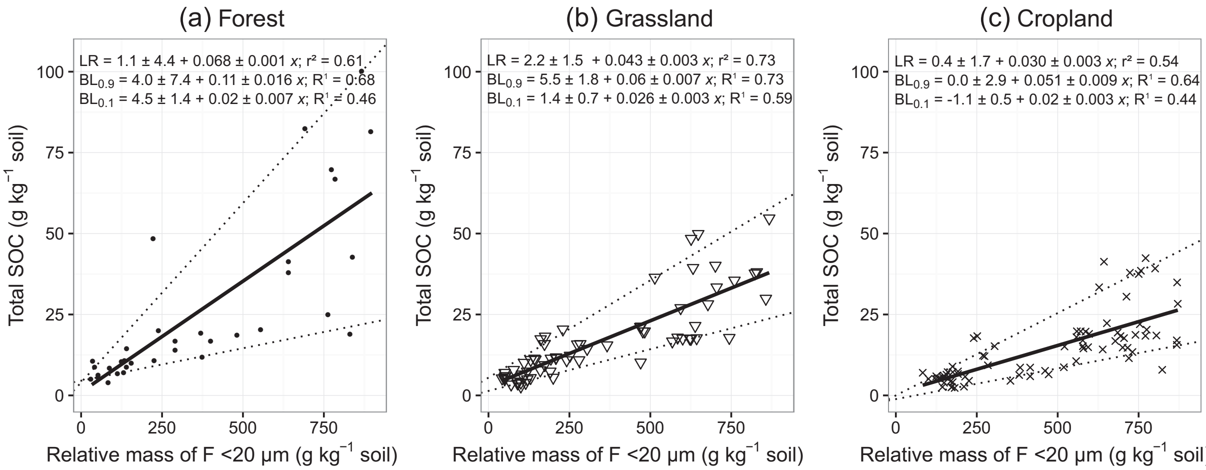

Fig. 4. Relationship between total SOC content (g C kg^{-1} soil) and relative mass of F < 20 µm fraction (g kg^{-1} soil) depending on land use (n = 34 for forests, n = 66 for grasslands, and n = 78 for croplands). The solid lines are the linear regression lines (LR). The dotted lines are the boundary lines (first decile, $BL_{0.1}$, ninth decile, $BL_{0.9}$).

unclear and varies with the dataset used. Hassink (1997) did not show any difference between tropical grassland and temperate grassland but that study used few soil samples. On the contrary Six et al. (2002b) and Feller and Beare (1997) showed that the silt and clay particles in tropical soils were less able to protect SOC than those in temperate soils, and explained this by the difference in clay mineralogy between

Table 2
SOC saturation deficit and SOC contents variations in control croplands and improved croplands or grasslands.

Reference Country, soil type	Land management	Clay + fine silt	SOC saturation	SOC saturation deficit[a]		SOC content change		
			C_{sat}	$C_{sat\text{-}def}$	$C_{sat\text{-}def\ LU}$	< 20 µm	> 20 µm	Total soil
		g kg^{-1} soil	g C kg^{-1} soil	g C kg^{-1} soil	g C kg^{-1} soil	g C kg^{-1} soil	g C kg^{-1} soil	g C kg^{-1} soil
(Feller et al., 1991)[§]	Peanut/millet[b]	123	6.5	3.0	1.1			
Senegal, Arenosol	Fallow in rotation	125	6.6	2.9	1.0	+ 0.2	+ 0.7	+ 09
	Fallow in rotation	152	8.1	4.2	1.9	+ 0.3	+ 0.3	+ 0.6
(Feller et al., 1991)[ß]	Maize[b]	133	7.0	3.2	1.2			
Togo, Ferralsol	NPK amendment	136	7.2	3.7	1.6	− 0.3	+ 0.2	− 0.2
	NPK amendment	156	8.3	3.6	1.2	+ 0.9	+ 0.2	+ 1.1
(Hien, 2004)[‡]	Sorgho[b]	182	9.6	8.4	5.6			
Burkina Faso, Acrisol	NPK amendment	211	11.1	9.3	6.1	+ 0.6	+ 0.1	+ 0.4
	NPK + manure	149	7.9	6.2	3.9	+ 0.5	+ 1.1	+ 1.6
	NPK + manure, high doses	176	9.3	6.3	3.7	+ 1.7	+ 3.2	+ 5.0
(Hien, 2004)[‡]	Sorgho[b]	172	9.2	7.5	4.9			
Burkina Faso, Acrisol	NPK amendment	173	9.2	7.6	4.9	+ 0.0	+ 0.2	+ 0.1
	Manure	176	9.3	7.4	4.8	+ 0.2	+ 1.3	+ 1.6
	NPK + Manure	169	9.0	6.8	4.2	+ 0.5	+ 1.3	+ 1.8
(Solomon et al., 2000)[ß]	Cultivated[b]	381	20.2	15.3	9.6			
Tanzania, Luvisol	Manure addition	305	16.2	3.4	− 1.2	+ 7.9	+ 0.9	+ 8.8
(Metay et al., 2007)[§]	Till + no cover crop[b]	586	31.0	18.2	9.4			
Brazil, Ferralsol	No till + cover crops	578	30.6	16.8	8.1	− 0.5	− 0.4	− 0.9
(de Freitas et al., 2000)[§]	Conventional Till[c]	586	31.0	14.1	5.2			
Brazil, Ferralsol	No tillage	559	29.6	13.3	4.9	− 0.7	+ 0.8	+ 0.1
(T. Razafimbelo et al.,	Sugar cane + fire[b]	740	39.2	24.5	13.4			
2006)[§] *Brazil, Ferralsol*	No fire + mulch	754	40.0	22.0	10.6	+ 3.3	+ 0.1	+ 3.3
(Grandière et al., 2007)[ß]	Conventional Till[b]	713	37.8	17.1	6.4			
Madagascar, Ferralsol	No till + mulch	723	38.3	13.7	2.9	+ 3.9	+ 3.5	+ 7.4
	No till + cover crop	757	40.1	12.9	1.6	+ 6.5	+ 1.5	+ 8.0
	No till + cover crop + legumes	750	39.8	14.0	2.7	+ 5.1	+ 1.9	+ 7.0
(Feller et al., 1991)[ß]	Market gardening[b]	805	42.7	25.8	13.7			
Guadeloupe, Vertisol	Grassland	823	43.6	13.5	8.6	+ 13.2	+ 6.1	+ 19.2
(Feller et al., 1991)[ß]	Market gardening[b]	720	38.2	28.4	17.6			
Martinique, Vertisol	Grassland	705	37.4	12.0	7.7	+ 15.6	+ 6.3	+ 22.0
(Larré-Larrouy et al., 2003)[‡]	Market gardening[b]	557	29.4	21.2	12.9			
Martinique, Vertisol	Grassland	626	33.2	9.8	6.0	+ 15.1	+ 22.7	+ 37.8
(Chevallier, 1999)	Bare soil[b]	731	38.7	29.4	18.5			
Martinique, Vertisol	Grassland[‡]	744	39.4	28.3	23.9	+ 1.8	+ 2.6	+ 4.4
	Grassland[§]	643	34.1	15.6	11.7	+ 9.2	+ 18.7	+ 27.9

Experiment duration: 3–9 years[§], 10–15 years[ß], > 15 years[‡].

[a] $C_{sat\text{-}defLU}$: SOC saturation deficit calculated using $BL_{0.9}$ calculated for each land use, for cropland $C_{satCROPLAND} = 0.038 \times F < 20\ \mu m$ and $C_{satGRASSLAND} = 0.047 \times F < 20\ \mu m$; C_{sat} values were expressed in g C in F < 20 µm kg^{-1} soil.

[b] Control.

tropical and temperate soils. Six et al. (2002b) showed that the tropical dataset they used was dominated by soil samples with 1:1 clays which have a lower specific surface area and, therefore, a lower capacity to bind and stabilize SOC than 2:1 clays, these latter being mainly found in temperate soils.

Although our dataset contained 2:1 clay soils, these soils did not show high SOC content in F < 20 μm and, therefore, had little influence on the SOC saturation line (Fig. 2b). This unexpected result was explained by the soil properties and the land use of the 2:1 clay-soils in our dataset. These 2:1 clay-soils were mainly cropped Vertisols. Vertisols are sensitive to dispersion and erosion, especially if they have high exchangeable sodium and magnesium content (Cook et al., 1992; Venkateswarlu, 1987). In cropped Vertisols, the biotic and fauna activities are too low to counteract this sensitivity to erosion by improving the soil structure (Shabtai et al., 2014). Long term cropping in Vertisols can, therefore, lead to SOC depletion through topsoil erosion, explaining the low level of SOC associated with F < 20 μm in the 2:1 clay-soils in our dataset.

As expected, forests soils, characterized by large C inputs and low soil disturbance, dominated the overall $BL_{0.9}$. The $BL_{0.9}$ calculated for forest (87 ± 12 g C kg^{-1} F < 20 μm) was slightly lower than the value (107 ± 8 g C kg^{-1} F < 20 μm) found by Feng et al. (2013) using a dataset including both temperate and tropical forests. The SOC content in F < 20 μm for the forest soils in our dataset showed a large dispersion, especially when F < 20 μm was higher than 650 g kg^{-1} soil. The highest values of SOC content could be explained by the volcanic nature of the parent material and the elevation of these sites (above 1000 m) leading to relatively low mean annual temperatures and reduced mineralization rates of organic matter (Solomon et al., 2002; Tchienkoua and Zech, 2004). Araujo et al. (2017) reported that the variability of SOC stocks in forest soils in tropical highlands was poorly explained by the fine particle content of these soils. In addition, primary forest, secondary forest, and dense tree plantations were grouped as a single forest category in our study. This clustering explained a part of the large variability of the SOC associated with the F < 20 μm and the total SOC content (Figs. 2 and 4), as conversion of primary forest to secondary forest or to plantations could result in a significant decline in the SOC stocks especially in the topsoil (Chiti et al., 2014; Don et al., 2011).

The $BL_{0.9}$ for grassland (47 ± 6 g C kg^{-1} F < 20 μm) was lower than the value (89 ± 5 g C kg^{-1} F < 20 μm) found by Feng et al. (2013) using a dataset including both temperate and tropical grassland. This difference could be explained by the clay type and grazing management. The dataset used by Feng et al. (2013) included 97 samples from grassland with 2:1 clays and 32 of these samples were over-saturated. In our study only 15 samples were soils with 2:1 clays, therefore the amount of stabilized SOC is expected to be lower than observed by Feng et al. (2013). In addition, our dataset included of abandoned pastures, fallows, or savannahs which were expected to receive low carbon inputs and, therefore store less SOC. The dispersion of SOC content for grasslands might be also due to differences in grazing management (grazing pressure, fertilizers inputs, and grass species). Grazing management directly impacts the SOC stocks through the reduction of above ground biomass, land cover and species diversity (Milne et al., 2015). Innovative grassland management and restoration of degraded grazed lands were reported to increase the SOC content (Akpa et al., 2016; Chaplot et al., 2016) which may increase the SOC associated with F < 20 μm. The samples closest to the global $BL_{0.9}$ were mainly found in Amazonia (Desjardins et al., 2004; Fujisaki, 2014), where it has been shown that SOC stocks in grasslands can be higher than SOC stocks in forests (Fujisaki et al., 2015; Stahl et al., 2016).

The $BL_{0.9}$ (38 ± 4 g C kg^{-1} F < 20 μm) calculated for cropland was slightly lower than the value (45 ± 5 g C kg^{-1} F < 20 μm) found by Feng et al. (2013) using a dataset including both temperate and tropical cropland. The SOC depletion in croplands compared to native vegetation has been widely reported for the tropics (Don et al., 2011; Powers et al., 2011). It is mostly caused by low carbon inputs and high soil disturbance. As for grasslands, differences in management that influence carbon inputs and soil disturbance could explain the large dispersion of the dataset values for croplands.

Apart from the textural control of SOC retention, there are other factors that could affect SOC stabilization (Zinn et al., 2007). For a New Zealand's dataset (Beare et al., 2014), the maximum stabilized SOC content was more closely correlated with extractable Al, surface area of fine particles, and soil pH than with fine particle mass alone. This is also especially true in tropical soils, where soils could contain iron and aluminum oxides and hydroxides which affect the soil structure, SOC dynamics and SOC accumulation (Barthès et al., 2008; Six et al., 2002b; Zinn et al., 2007). Oxides and hydroxides have high specific surface areas and can adsorb dissolved SOC but they also can flocculate, reducing the surface available for SOC adsorption (Six et al., 2002a). These opposing effects are difficult to prioritize. Unfortunately, extractable Al and Fe contents were not routinely reported in the studies and therefore were not included in our study. Improving datasets by systematically collecting information on such predictors is required to understand and quantify accurately the SOC stabilization mechanisms in tropical soils.

4.2. SOC changes in SOC associated with F < 20 μm with changes in cropland management

In croplands, the mean SOC content was half the SOC saturation value derived from the $BL_{0.9}$ for the whole dataset, suggesting that there is a large potential for SOC accumulation, even for soils with < 250 g of F < 20 μm kg^{-1} soil (Fig. 2). Overall, the comparison between the linear regression lines and the upper boundary lines showed that the SOC saturation deficit in F < 20 μm increases with the relative mass of F < 20 μm in the soil. Therefore, the potential for SOC accumulation was potentially larger for clayey soils than that for sandy soils (Fig. 1). However, the change in the SOC content in F < 20 μm after a change in agricultural management practices (e.g. manure inputs, crop residues return, no-tillage, rotation with fallow or pasture period, and mineral amendment) was, however, reported for only 9 cases (Table 2). While the conversion of annual crops plots into grassland also appeared to be an effective way of increasing the SOC content in F < 20 μm (Table 2), this was reported in only 4 cases. It is difficult to elaborate with so few case studies but these results agree with other studies that reported large increases of SOC stocks in the tropics when cropland is converted to grassland (Don et al., 2011).

The amount of C inputs probably explained the difference observed between the ability of the “improved” practices to fill the SOC saturation deficit and the SOC saturation level, as C inputs are a major determinant of SOC dynamic (Virto et al., 2012). Nevertheless, as the experimentation duration ranged from 6 to 11 years in some studies from Table 2 dealing with agricultural management changes (de Freitas et al., 2000; Grandière et al., 2007; Metay et al., 2007; Razafimbelo et al., 2006; Solomon et al., 2000), the SOC accumulation may continue before reaching an equilibrium state, since SOC accumulation is partly controlled by experiment duration (Minasny et al., 2017).

More interestingly, the dataset we collected revealed that improving management of cultivated soils or converting to grassland did not make these soils reach SOC saturation within several years or even decades. We, therefore, adapted the concept of “effective stabilization capacity” proposed by Stewart et al. (2007). We defined an upper limit of SOC storage which depends not only on the physical and chemical properties of the soils but also from the levels of SOC observed for a given land use. This reduces the SOC target level to a level that is achievable for cropland. Using the boundary line ($BL_{0.9}$) calculated for cropland or grassland instead of the $BL_{0.9}$ obtained from the whole dataset drastically reduced SOC saturation deficit, and allow overcoming the effect of outliers data in forest soils on SOC saturation curve (Fig. 2). This means

that the soil under annual crops with improved practices can reach this SOC saturation level calculated specifically for cropland. $C_{sat\text{-}defCROPLAND}$ could, therefore, provide a more realistic estimation of the SOC saturation deficit in tropical croplands and could be used as a proxy for an achievable objective for SOC saturation in agricultural lands.

However, the cases presented in Table 2 were synchronic studies, where there are errors associated with the difficulty of finding situations with the same initial soil conditions, i.e. with well-known previous land uses and identical management history, which often leads to over or underestimating the changes in SOC content (Costa Junior et al., 2013). Diachronic studies are needed to properly test whether soils that have not reached their SOC storage limit in the fine fractions, really do have the potential to store additional carbon.

4.3. The contribution of SOC associated with coarser soil fractions

The SOC associated with coarser fractions is not included in the SOC stabilization concept, because the SOC turnover rate is much higher in the coarse than in the fine fractions (Baldock and Skjemstad, 2000), even where the OC in the coarser fractions can be protected against microbial degradation through spatial inaccessibility or reduction of substrate diffusion due to physical occlusion in soil aggregates (Baldock and Skjemstad, 2000; Chevallier et al., 2004). However, the coarser soil fractions, mostly comprising particulate organic matter, are an important part of the total SOC content in tropical soils. Despite the large data dispersion, 25.9 ± 15.1% of the total SOC content was found in F > 50 μm (Table 1). The amounts of SOC associated with these coarse fractions were independent of the soil texture (Fig. 3), which was also shown by Zinn et al. (2007) in soils of Brazilian Cerrados. Consequently this SOC pool associated with F > 50 μm was significantly higher in sandy soils (41.3 ± 15.4% of total SOC content) than in other soils. This is remarkably similar to the data observed in sandy soils in Zinn et al. (2007). The SOC content in F > 50 μm was significantly higher in forest and in grassland than in cropland, but the relative contribution of this pool to total SOC content was, surprisingly, not significantly different between the three land uses when all soil groups were merged (Table 1). Gregorich et al. (2006) found, in a meta-analysis, that the proportion of this pool in soil was 27.9 ± 11.3% in forest, 20.8 ± 10.9% in grassland, and 21.6 ± 15.6% in cropland.

Moreover, the increased SOC content after a change in the cropland management was also found in the coarser fraction (here the F > 20 μm) (Table 2). This highlights the contribution of the SOC pool in the coarser fractions to the increase in total SOC content after increasing the amount of biomass added to the soil (Feller et al., 1991).

4.4. SOC storage potential in the tropics

The SOC storage potential is usually related to the SOC stabilized in F < 20 μm. However, to enlarge the basis for estimating SOC storage potential we established a relationship between mass of F < 20 μm and total SOC content (Fig. 4) as in Feller et al. (2001), although the r^2 values were lower than with SOC contents in F < 20 μm (Fig. 2). As the mass of the F < 20 μm can be obtained through a classical texture analysis and SOC in particle-size fractions data will likely remain scare due to analytical constraints, the positive relationship between total SOC content and the relative mass of F < 20 μm may be used as an indicator for the SOC storage potential. This is interesting as data are still needed to define properly the SOC storage potential of agricultural lands. Indeed, the $BL_{0.9}$ calculated in the tropics needs further assessment, since the soil, climate, and spatial representativeness of the dataset were unbalanced. Studies dealing with SOC distribution between particle-size fractions were common for West Africa and Latin America, but scarce for South East Asia, the Indian subcontinent and Southern Africa. Some tropical soils, such as Cambisols were underrepresented in our dataset, although they are widespread in the Indian subcontinent, representing 39% of Indian soils according to (Bhattacharyya et al., 2013), and are widely used as cropland. The levels of SOC saturation defined in this study may change in the future thanks to further assessments, especially in Asia and in the Indian subcontinent.

Our study highlights the need to consider a $BL_{0.9}$ defined with croplands observations instead of mixed land-use when calculating SOC storage potential for croplands. Using $BL_{0.9}$ defined with croplands observations avoid overestimation of SOC storage potential by using the SOC saturation deficit as calculated for temperate soils (Angers et al., 2011; Wiesmeier et al., 2014). For instance, Ferralsols, the most common soil type in our dataset, seem to have a large SOC storage potential. They cover large areas in the tropics, 750 million ha (IUSS Working Group WRB, 2015), and they are depleted in SOC under croplands by comparison with forest or pristine vegetation. They could, therefore, potentially store a large amount of SOC, filling the SOC saturation deficit as shown in Fig. 1a. However, as discussed in Barré et al. (2017), our results indicate that caution is required, especially if we expect to increase soil carbon stocks in croplands with changes in the management practices rather than land use. The first caution is that the $BL_{0.9}$ calculated for all land uses taken together provides a target that is very unlikely to be reached in cultivated soils (Table 2). The second is that “improvements” in land management practices did not always increase the SOC content. One reason could be that, in some cases, the carbon inputs were insufficient to increase the SOC content. This is a major constraint in tropical agroecosystems, where competition for organic materials is a barrier to the adoption of improved management practices (Giller et al., 2009; Vanlauwe et al., 2014). However, one can hope that the levels of SOC saturation defined in this study may change in the future thanks to further assessments of innovative cropland management.

Studies that estimate SOC storage potential using specific SOC saturation lines at national scale (Beare et al., 2014; McNally et al., 2017) would be welcome in the tropics. Increasing SOC contents and stocks have two main objectives food security and climate (Minasny et al., 2017). The different SOC pools in the fine and coarse fractions do not have the same impact on the soil properties for food security or climate issues. Soil fine fractions are recognized to stabilize OC and consequently to play a role in climate mitigation, but Wood et al. (2016) did not observe a positive effect of stabilized SOC on crop yield production. This indicates that we need to revise the view that stabilization of organic matter improves food security. Arenosols in West Africa have low SOC storage potential in F < 20 μm (Fig. 1a), and improved management practices had no clear effect on SOC in F < 20 μm for these soils (Table 2). In addition, the very small accumulation in SOC in F < 20 μm is likely to be difficult to measure. Thus, implementing practices aiming to increase stabilized SOC content to contribute to climate change mitigation is not the priority for Arenosols. However, some studies have advised regular C inputs to maintain a sufficient level of soil fertility (Freschet et al., 2008; Janssen, 2011) to provide nutrients for crop growth and to avoid the degradation of the soil quality caused by continuous cultivation (Mapfumo et al., 2007; Ouédraogo et al., 2001; Bationo et al., 2007; Rawls et al., 2003). Moreover, increasing SOC associated with F > 50 μm through increased organic inputs would be also worthwhile for agriculture adaptation to climate change, increasing the water holding capacity and drought preparedness (Smith and Wollenberg, 2012).

5. Conclusions

Our results are a warning against overestimating SOC storage potential by using the SOC saturation deficit as calculated for temperate soils, and provided a specific estimation of the SOC saturation levels in the tropical soils thanks to boundary line analysis of SOC contained in fine particle-size fractions (F < 20 μm). We found a large C storage potential in croplands. However we found that changing agricultural practices in croplands did not allow soils to reach SOC saturation line, especially that defined from the whole dataset.

We strongly recommend using saturation line defined with data from a single land-use (e.g. grassland or cropland) instead of mixed land-use as a more realistic target when calculating SOC storage potential.

We underlined the lack of data in some tropical regions, especially in Asia and in the Indian subcontinent. The levels of SOC saturation defined in this study may change in the future thanks to further assessments. Long-term studies dealing with innovative agricultural management like agroforestry, use of cover crops, or conservation agriculture may define new levels for cropland saturation line. The positive relationship between total SOC content and the relative mass of F < 20 μm may be used as an indicator for the SOC storage potential.

In the framework of international initiatives that promote soil carbon storage for food security and climate change mitigation (e.g. 4P1000, UNCCD), methodology based on boundary lines of SOC retention could help policy makers and farmers to evaluate the opportunity of store carbon efficiently, in the light of adapted and realistic targets. Work is still needed to clarify the impact of SOC stabilization on crop yields and the link between SOC saturation and food security. This is essential for proposing soil organic matter management methods in tropical environments where competition for organic material is often critical.

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